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ABSTRACT

Determination of the stability regime is a basic approach in any investigation of atmospheric turbulence. The establishment of stability criteria in the boundary layer is usually accomplished by use of the nondimensional Richardson number. The computation of accurate Richardson numbers is shown to be adversely affected by a number of factors including the choice of vertical gradients, the terrain, spacing of instruments, and heterogeneous profiles of wind and temperature.

CONTENTS

INTRODUCTION

The analysis of turbulent processes in the first few meters of the atmosphere is usually based upon some scheme for defining the stability regime in operation at the time the experimental data are collected. The regimes may be classified by any number of methods as long as the classification system yields the desired results. The most common classifier of stability is the Richardson number, which is quite adequate if certain precautions are observed in its calculation. To use the Richardson number effectively as an identifier of the stability regime, it is necessary to understand the turbulent processes within the surface boundary layer.

Since the numerical calculation of the Richardson number is highly dependent upon the vertical gradients of wind velocity and temperature, proper evaluation of these parameters is vital in terms of whether the data are representative or have been biased by horizontal advection or the presence of local terrain effects that lead to unsteady-state flow.

Compensation for nonhomogeneous processes in the boundary layer can be difficult, if all the contributing factors cannot be identified or isolated. Some of the nonuniform effects on the accurate determination of the Richardson number have been investigated with respect to unsteady-state flow in the surface boundary layer. The results indicate that a meaningful Richardson number may be computed with confidence using heterogeneous experimental data.

The purposes of this report are to discuss (1) the Richardson number, including computation problems which arise using actual data, (2) nonequilibrium effects on profile gradients, and (3) measurement requirements of vertical gradients near the surface and to present results obtained by the author from data collected at White Sands Missile Range and treated in view of the limitations presented in (1) and (2) above.

DISCUSSION

The Richardson number, a nondimensional parameter possessing the characteristics of dynamic similarity according to Batchelor (1953), is the accepted stability indicator in most studies concerning atmospheric turbulence. Richardson (1920, 1925), while investigating the effects of gravity on the suppression of turbulence, derived a ratio of work done against gravitational stability to energy transformed from mean to turbulent motion. It was asserted that a motion which was slightly turbulent would remain so if the ratio were less than one and would subside if the ratio were greater than one.

Richardson's criteria was most simply described by Brunt (1941) to be

$$Ri = \frac{g}{\overline{T}} \frac{\frac{\partial \overline{T}}{\partial z} + \Gamma}{\left(\frac{\partial \overline{V}}{\partial z}\right)^2} \qquad (1)$$

where g is the acceleration due to gravity, $\overline{T}$ is mean temperature (^{o}K) at the level of interest, and the gradients, and Γ the dry adiabatic lapse rate. Richardson's original assumption that the exchange coefficients for heat and momentum were equal has been shown to be invalid, and Richardson's original ratio is now taken to be the flux form of the Richardson number, R_f, which according to Ellison (1957) may be expressed as

$$R_f = Ri \frac{K_H}{K_M} = \frac{K_H}{K_M} \frac{g \frac{\partial \overline{\rho}}{\partial z}}{\left(\frac{\partial \overline{V}}{\partial z}\right)^2} \tag{2}$$

where the exchange coefficients for heat and momentum are defined as

$$K_H = \frac{\overline{T' w'}}{\frac{\partial \overline{T}}{\partial z}} \tag{3}$$

and

$$K_M = \frac{\overline{u' w'}}{\frac{\partial \overline{V}}{\partial z}} \tag{4}$$

Ellison also suggests that turbulence subsides at a value of R_f less than one, and that the critical value is approached as a limit, under stable atmospheric conditions, such that

$$R_f \text{ crit.} = Ri \frac{K_H}{K_M} = g \frac{\overline{\rho' w'}}{\overline{\rho} u_*^4} K_M \tag{5}$$

where K_M is independent of height. Hence R_f as well as Ri has a critical value. The critical value of R_f is approximately 0.15 and according to McVehil (1962), critical Ri lies between 0.14 and 0.22, considerably less than Richardson's original estimate of 1.0. Thus, as Ri approaches a critical limit, the ratio K_H/K_M must decrease proportionally.

Experimental evidence based upon stationary conditions indicates that K_H/K_M is approximately one in forced convection; but the actual values for all stability conditions are still undetermined experimentally. Estimates are, depending upon the stability regime, from 0.70 (Senderikhima, 1961) to 1.6 (Ellison, 1957) with a geometric mean value of 1.3 in unstable conditions.

The basis of many hypotheses concerning the shape of the wind and temperature profiles in the boundary layer under diabatic conditions is the Richardson number. These include the independently derived models of Laikhtman (1944) and Deacon (1949), usually written as

$$\frac{\partial \overline{V}}{\partial z} = Az^{-\beta} \qquad (6)$$

where β is a shape parameter and is a function of the Richardson number. It was originally assumed that β was independent of height and thus a unique parameter, but Davidson and Barad (1956) and later Lettau (1962) showed this assumption to be in error.

From the similarity theory of Monin and Obukhov (1954) it can be shown that the Richardson number is a unique function of z/L', an arbitrary gradient length ratio defined by Panofsky, Blackadar and McVehil (1960) where

$$\frac{z}{L'} = SRi \qquad (7)$$

with S being defined as a nondimensional logarithmic wind shear. At least eight diabatic boundary layer profile models based upon the work of Monin and Obukhov and Eq. (7) have been developed as a function of the Richardson number.

The Richardson number also performs an important function in delineating the transition between forced and free convection which occurs at approximately $Ri = -0.03$. Priestley (1955) suggested that the transition was quite sharp. In a later paper Priestley (1959) found that the transition was rather gradual, as also determined by Panofsky, Blackadar, and McVehil (1960). From the theory of free convection it can be shown that the transition zone lies in the stability range $-0.02 > Ri > -0.05$ and that a junction height can be determined at $z/L = Ri = -0.03$, where L is the Monin-Obukhov scaling height.

Another characteristic of the Richardson number is a rather smooth trend toward larger absolute values with increasing height above the surface. Lettau and Davidson (1957) list values of Ri from 100 - 2000 meters above the surface for three stability classes, a contour number and the Deacon number (Table I). In a discussion of diabatic surface layer models, Lettau (1962) and Davidson and Barad (1956) stress the dependence of flow near the ground on flow conditions at greater heights; thus the tendency of the Richardson number to increase in absolute magnitude with height and especially the rate at which Ri increases with height will have considerable bearing on the wind profile shape in the boundary layer.

Height (Meters)	Extreme Lapse			Neutral			Extreme Inversion		
	Ri	α	β	Ri	α	β	Ri	α	β
2000	13.7	-0.50	0.07	2.9	-1.30	0.99	20.0	-1.10	2.10
1750	11.3	-0.50	0.19	2.7	-1.30	0.40	13.7	-1.10	2.10
1500	10.3	-0.51	0.45	2.4	-1.30	0.09	1.2	-1.10	2.10
1250	6.8	-0.45	0.81	2.5	-0.66	0.04	3.3	-1.10	1.70
1000	2.9	-0.40	0.00	2.4	-0.66	0.07	1.3	-1.10	0.90
800	12.6	-0.24	-1.63	3.4	-0.40	0.18	1.0	-0.26	0.00
700	21.5	-0.18	-1.23	2.8	-0.30	1.05	2.4	-0.26	-1.18
600	28.5	-0.15	-1.74	5.2	-0.22	1.05	4.1	0.00	-0.94
500	54.7	-0.07	0.00	3.1	-0.14	1.05	4.1	0.00	-0.71
400	5.3	0.07	0.10	1.6	0.00	0.61	4.0	0.00	0.00
300	3.4	0.08	0.57	0.9	0.15	0.61	4.3	0.00	0.84
200	1.3	0.08	0.54	0.6	0.17	0.16	3.1	0.16	0.65
100	0.7	0.07	0.40	0.7	0.14	0.07	3.9	0.19	0.45

Table I. Local values of Richardson Number, Ri, Profile contour Number, α, and Deacon Number, β, at indicated heights from class averages of free-air potential temperature and wind component data. (After Lettau)

METHODS FOR DETERMINING THE RICHARDSON NUMBER

The Richardson number is usually computed by use of Eq. (1) in the form

$$\mathrm{Ri} = \frac{g}{\overline{T}} \frac{\left(\frac{\partial \overline{T}}{\partial z} + \Gamma\right)}{\left(\frac{\partial \overline{V}}{\partial z}\right)^2} . \tag{8}$$

To facilitate computation, Eq. (1) can be restated as

$$\mathrm{Ri} = \frac{g}{\Theta} \frac{\Delta\overline{\Theta}}{(\Delta \overline{V})^2} z \Delta \ln z \tag{9}$$

and

$$\mathrm{Ri} = \frac{g}{\overline{T}} \frac{\Delta\overline{T} + \Gamma}{(\Delta \overline{V})^2} z \Delta \ln z \tag{10}$$

assuming "geometric progression" spacing of the instruments and finite difference determination of the gradients.

Lettau and Davidson (1957) define the Richardson number by

$$Ri = g(\ln\Theta)' / \overline{V}'^2 \tag{11}$$

with the primes denoting partial differentiation with respect to height. Near the surface this may be expressed as

$$Ri = g\ \Theta'/T_M\ \overline{V}'^2 \tag{12}$$

where T_M is the average temperature of the layer under consideration. Kutzback (1961) defines Ri as

$$Ri = g\Delta z \frac{\Delta\Theta}{T_M(\Delta\overline{V})^2} \tag{13}$$

A more simplified version of Ri is the "Poor Man's Richardson Number," a stability ratio which is used to define the stability regime and is given by

$$S.R. = \frac{\overline{T}_3 - \overline{T}_1}{V_2^2} \tag{14}$$

or

$$S.R. = \frac{\overline{T}_2 - \overline{T}_1}{V_1^2} \tag{15}$$

with the subscripts denoting the instrument level of measurement. The stability ratio has been used extensively by Deacon (1949) and Lettau and Davidson (1957).

Thus, it might be said that the determination of the Richardson number is rather arbitrary. Certainly, it depends upon the chosen computation method. All of the methods give similar results, with sampling time and interpretation being the most difficult problems. To solve the sampling problem, one should average over times commensurate with best estimates of the vertical transfer processes. According to Van der Hoven (1957), from the analysis of wind spectra, this estimate is approximately one hour's data; the estimate is further verified by results presented by Lumley and Panofsky (1964), which indicate that there is a gap in the spectrum at a period of one hour, separating the micro- and mesometeorological process scales in the atmosphere, indicative that approximately one hour of data will constitute a quasi-stationary microscale ensemble representative of prevailing conditions. If this is so, then boundary layer processes can be defined with some accuracy if hour samples of wind and temperature profiles are available.

However, shorter sampling periods may be used if prevailing synoptic and diurnal conditions are sufficiently stable to insure some stationarity, i.e., no cloud shadows are passing over the instrumentation site or measurements are not taken through sunrise or sunset; hence, two or more stability regimes are not combined in one data sample. According to Swinbank (1964) a cloud passing between the sun and an instrumentation array can cause nonequilibrium conditions that take up to 15 minutes or more to stabilize, owing to the change in heat flux from the abrupt drop in insolation. Thus, it can be seen that when many non-stationary processes are affecting the data, the measurement or evaluation of any micrometeorological parameter including the Richardson number becomes quite uncertain.

NONEQUILIBRIUM EFFECTS ON PROFILE GRADIENTS

Richardson numbers are dependent upon the vertical gradients of velocity and potential temperature. Since Ri is a ratio of work done to energy transformed from mean to turbulent flow, it is apparent that the definition of a mean is quite critical for accurate determination of the prevailing stability regime. To describe the mean accurately requires homogeneity of flow, i.e., defined as a gaussian and stochastic distribution of the variable.

For instance, a Richardson number that is to be used by Similarity Theory must be properly determined or it will lead to huge errors in the gradient and scale lengths, the normalized shear, the heat and momentum fluxes and the shearing stress. Since Similarity Theory requires that a system be stationary in both time and space, trends and heterogeneous flow resulting from terrain effects must be compensated for or they will tend to nullify any results obtained from application of the theory. In fact, most investigators of micrometeorological wind and temperature profiles assume homogeneity, or work over areas where they feel that the fetch is of sufficient distance upwind to assure homogeneous flow. However, fetch is only one factor to consider in assuming homogeneous flow. Local advection may cause nonhomogeneity even if a system is steady with time. Good indicators that advection is occurring are nonequilibrium vertical gradients of wind shear and potential temperature.

Philip (1959) derived a theory of local advection based upon the conjugate laws and near-neutral conditions that indicated that changes of heat and moisture fluxes at the air-earth interface led to vertical gradients of the fluxes that were variable with height for some distance downwind. Dyer and Pruitt (1962), while comparing eddy-flux determinations at a height of 4 meters and a fetch of 130 meters over an irrigated field surrounded by arid areas, found horizontal gradients of temperature and humidity of considerable magnitude between the surface and 4 meters. Applying Philip's method, Dyer (1963) found that adjustment of gradients with distance from a leading edge (abrupt change of flux) or with time required more distance or time than heretofore suspected. Until recently, the value of the fetch-height ratio for equilibrium flow was of the order of 50 to 75:1. Dyer (1963) reported that the fetch-height ratio was as large as 530 for 90 percent adjustment at a height of 50 meters with a time as long as 86 minutes, indicating that homogeneous fetches as much as 10 times greater than had been used previously are needed for equilibrium conditions.

Philip (1959) and Dyer (1963) assumed no change in surface roughness in conjunction with the change in heat and moisture flux. The effect of change in roughness was predicted by Ellison (1957) and investigated by Elliott

(1958a, 1958b). Elliott's theory of internal boundaries was extended by Panofsky and Townsend (1964). Basically, the hypothesis assumes that air flowing across a surface and encountering an abrupt change in roughness will experience an acceleration in the layer next to the surface causing an internal boundary to form. The flow beneath the boundary will possess the characteristics of the new surface while the flow above the boundary will exhibit the characteristics of the terrain upstream of the discontinuity. The slope of the boundary appears to be 1:10 with a fairly sharp interface separating the masses of air influenced by the two types of terrain. Elliott (1958b) found that the basic relationship can be expressed approximately by

$$h = 0.86x^{0.8}z_o^{0.2} \tag{16}$$

where h is the height of the internal boundary, x the distance downwind from the leading edge discontinuity, and z_0 is the roughness length. In the above hypothesis, Elliott (1958b) states that unstable conditions lead to an increase in the height of the boundary and stable conditions lead to a decrease in the height of the boundary, relative to the adiabatic case. It is clear, then, that surface discontinuities and advection can lead to the establishment of heterogeneous horizontal gradients and nonstationary time series that result in variable vertical gradients. Therefore, data derived from observations under such conditions for determining the Richardson number and indirectly the parameters necessary for application of diabatic profile theory, without considering fetch and flux effects, should be used with caution.

THE MEASUREMENT OF VERTICAL GRADIENTS NEAR THE SURFACE

Vertical gradients are usually calculated by a finite-difference approximation, where the level of interest is midway between the levels where measurements are made or at the geometric mean of the two levels. According to Bernstein and Young (1962), considerable error may occur if the gradient varies with height, as in the case of local advection or the internal boundary.

Gradients such as   or   are usually measured by assuming a finite difference such that

$$\frac{\Delta T}{\Delta z} = \frac{\bar{T}_2 - \bar{T}_1}{z_2 - z_1} \tag{17}$$

is a good approximation. The levels z_1 and z_2 are generally selected so that the level of interest falls at the midpoint. Since the mean temperature and mean wind are not usually linear functions of height but are more likely to be logarithmic or exponential, it is clear that a linear interpolation may be valid only under neutral conditions.

To compensate for nonlinear gradients, Bernstein and Young (1962) developed a technique for determining the heights at which instruments must be mounted to provide gradient measurements at the height of interest. Also, correction factors are provided to obtain the gradient at any height from measurements at any two heights, provided the profile shape is known. Thus, the nonlinearity of profile may be compensated for, but not the nonstationarity.

If two sensors are located equal distances above and below the level of interest; the value obtained for the gradient is too large and is actually the value for some level below the level of interest. The error will tend to increase as the separation between the sensors increases (Bernstein and Young, 1962). Errors can range from 1 percent to 7 percent when the separation is half as great as the height above ground of the level of interest or from 3 percent to 34 percent for a separation distance equal to the height of interest or from 10 percent to 13 percent for a separation 1½ times the height of the level of interest.

RICHARDSON NUMBER USING WSMR DATA

Same Richardson numbers and stability ratios computed from data obtained on the 62-meter research tower at White Sands Missile Range are presented in Tables II, III and IV. The data used to evaluate Ri is considered to be heterogeneous owing to the prevailing conditions at the time of observation. The data for 0234-0333 MST, 26 January 1962, were taken while a nocturnal drainage wind was occurring and modifying the prevailing flow. The data for 1600-1659 MST, 7 May 1962, were observed during a highly unstable period. It appears that the internal boundary at the tower site (Hansen and Hansen, 1965) was completely masked by free convection effects overriding the mechanical processes generating the internal boundary. The data for 1330-1429 MST, 5 February 1962, were observed with the mean flow across heterogeneous terrain. Wind and potential temperature profiles for the three periods are presented in Figures 1, 2 and 3.

Height (Meters)	Ri (1)	Ri (9)	Ri (12)	Ri (13)	S.R. (14)	S.R. (15)
2.95	0.022	0.011	0.015	0.011	0.0031	0.0035
5.33	0.023	0.023	0.016	0.023	0.0022	0.0024
9.65	0.045	0.027	0.010	0.027	0.0020	0.0022
15.92	0.018	0.016	0.008	0.016	0.0009	0.0010
22.10	0.033	0.026	0.010	0.026	0.0006	0.0006
28.22	0.027	0.021	0.014	0.021	0.0004	0.0004
34.36	0.037	0.073	0.061	0.074	0.0004	0.0004
43.21	0.042	0.023	0.005	0.024	0.0004	0.0004
55.50	0.054	0.039	0.013	0.038	0.0003	0.0003

Table II. Richardson numbers and stability ratios for 0234-0333 MST, 26 January 1962. The numbers in parentheses are the equation numbers used. Nocturnal drainage wind conditions.

Height (Meters)	Ri (1)	Ri (9)	Ri (12)	Ri (13)	S.R. (14)	S.R. (15)
2.95	-0.030	-0.069	-0.165	-0.070	-0.0188	-0.0210
5.33	-0.090	-0.086	-0.066	-0.087	-0.0108	-0.0122
9.65	-0.101	-0.123	-0.037	-0.125	-0.0085	-0.0095
15.92	-0.264	-0.168	-0.086	-0.174	-0.0036	-0.0038
22.10	-0.212	-0.234	-0.124	-0.233	-0.0021	-0.0022
28.22	-0.254	-0.194	-0.194	-0.195	-0.0011	-0.0011
34.36	-0.342	-0.257	-0.176	-0.262	-0.0008	-0.0008
43.21	-0.332	-0.221	-0.038	-0.225	-0.0009	-0.0010
55.50	-0.401	-0.461	-0.103	-0.454	-0.0009	-0.0009

Table III. Richardson numbers and stability ratios for 1600-1659 MST, 7 May 1962. The numbers in parentheses are the equation numbers used. Highly unstable daytime conditions.

Height (Meters)	Ri (1)	Ri (9)	Ri (12)	Ri (13)	S.R. (14)	S.R. (15)
2.95	-0.288	-0.490	-0.449	-0.499	-0.0494	-0.0563
5.33	-0.318	-0.403	-0.231	-0.408	-0.0264	-0.0290
9.65	-0.334	-0.477	-0.127	-0.486	-0.0181	-0.0211
15.92	-1.217	-0.976	-0.394	-1.007	-0.0074	-0.0077
22.10	-2.156	-1.795	-0.902	-1.784	-0.0045	-0.0047
28.22	-0.766	-0.700	-0.557	-0.704	-0.0022	-0.0022
34.36	-4.158	-2.331	-2.038	-2.379	-0.0026	-0.0027
43.21	-5.040	-4.264	-0.505	-4.333	-0.0030	-0.0041
55.50	-4.684	-4.447	-1.143	-4.402	-0.0020	-0.0020

Table IV. Richardson numbers and stability ratios for 1330-1429 MST, 5 February 1962. The numbers in parentheses are the equation numbers used. Flow across nonuniform terrain.

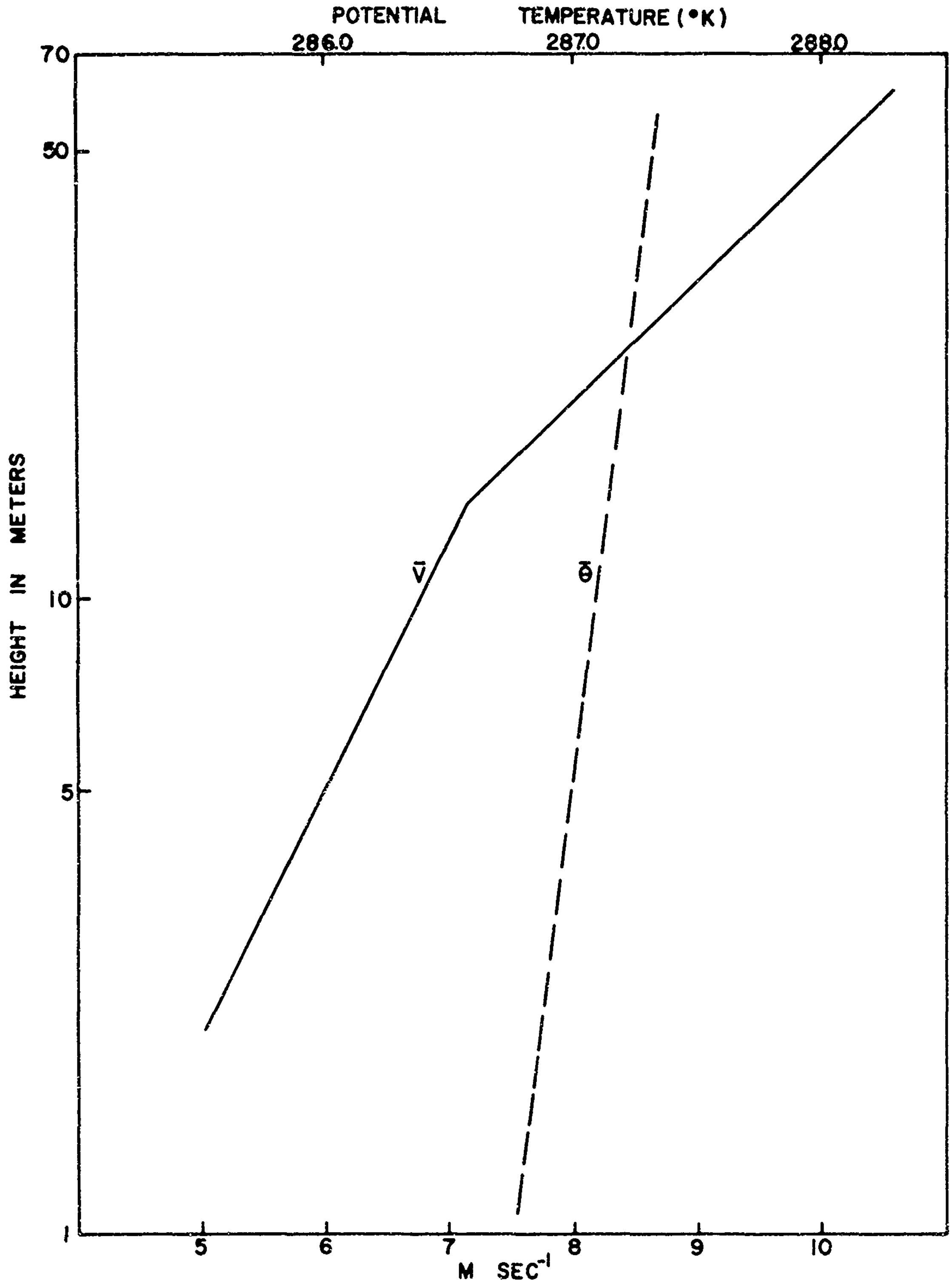

FIGURE 1: WIND AND POTENTIAL TEMPERATURE PROFILES FOR 0234-0333 MST, 26 JANUARY 1962.

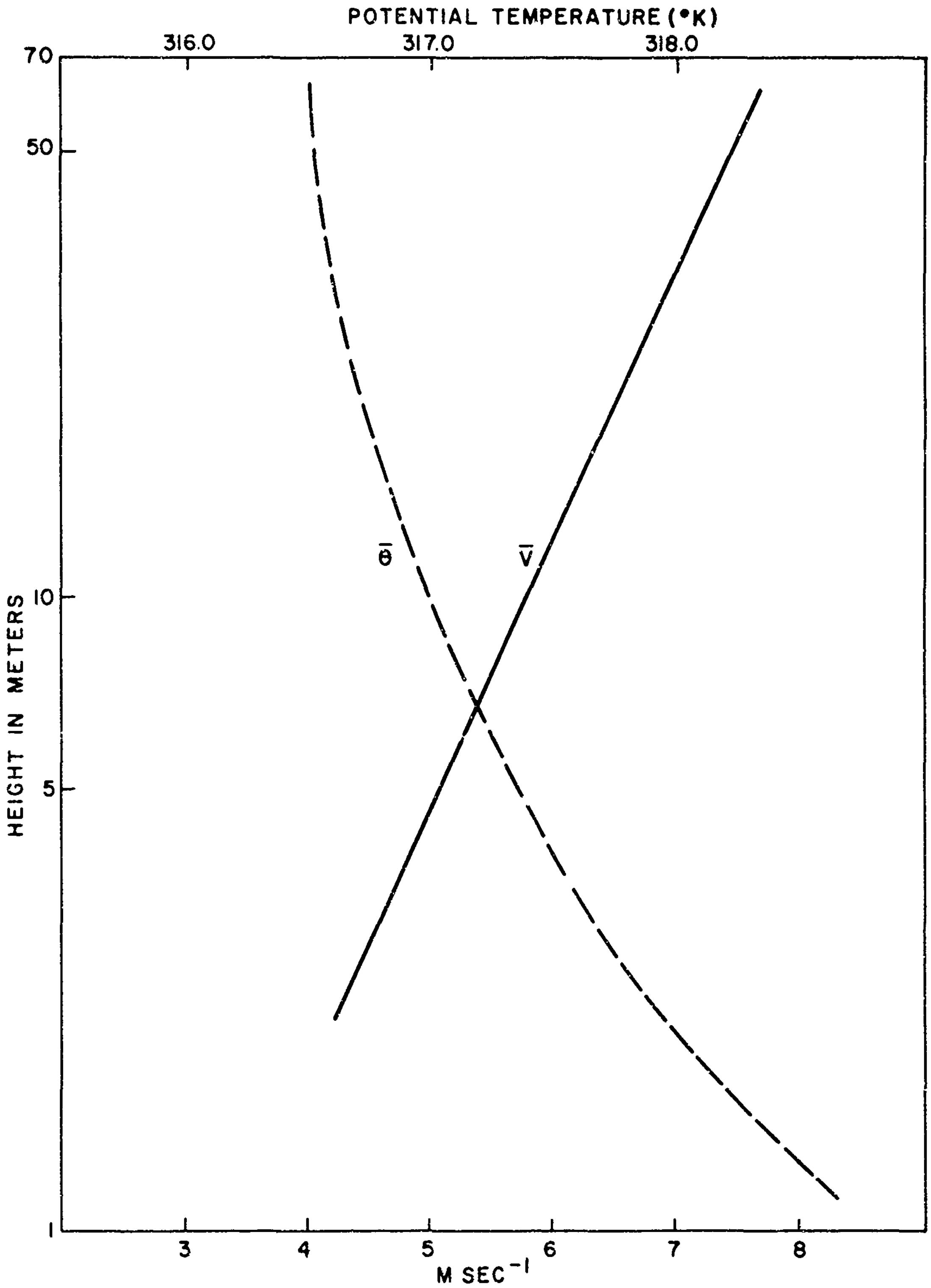

FIGURE 2: WIND AND POTENTIAL TEMPERATURE PROFILES FOR 1600-1659 MST, 7 MAY 1962

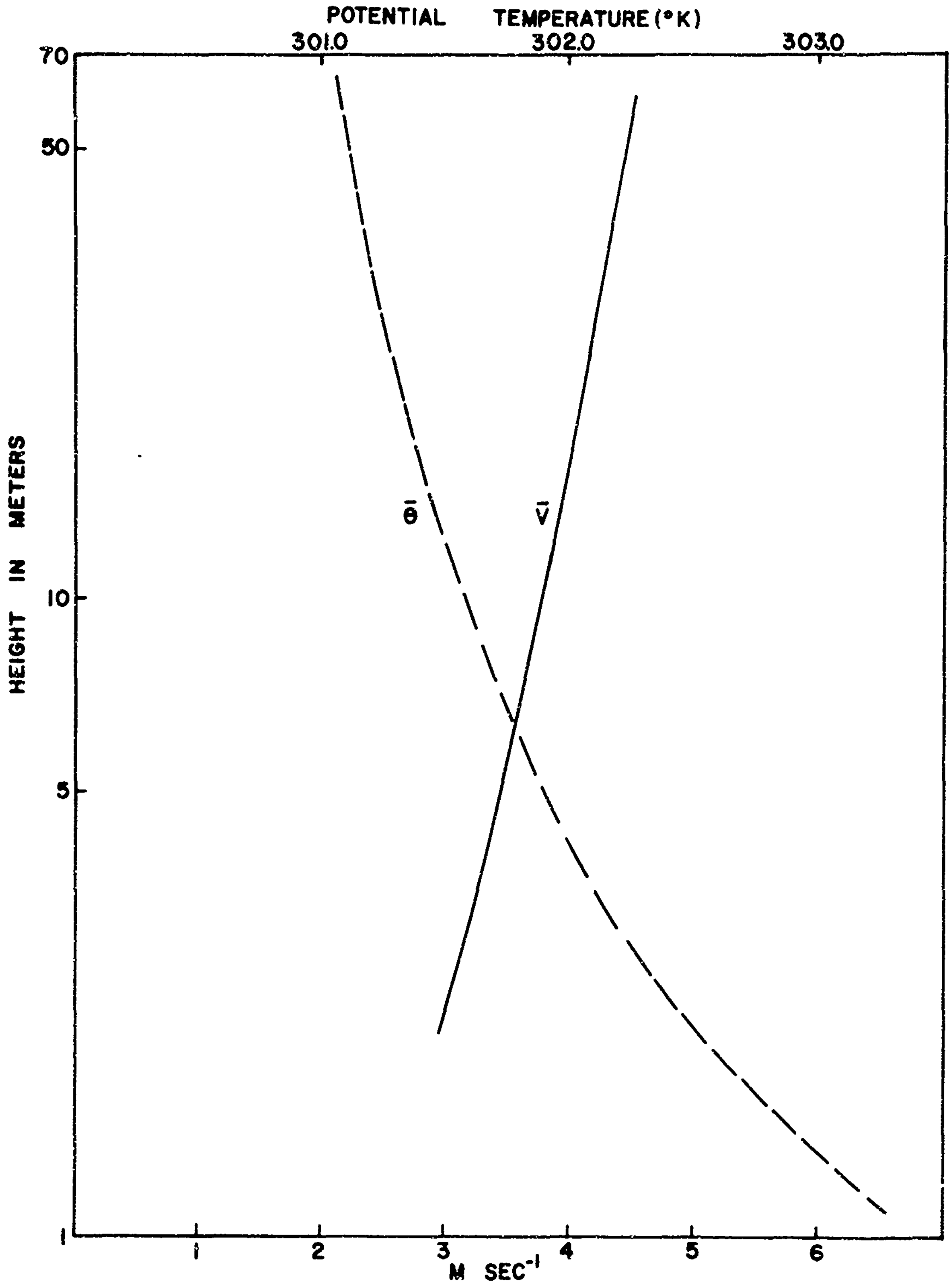

FIGURE 3: WIND AND POTENTIAL TEMPERATURE PROFILES FOR 1330-1429 MST, 5 FEBRUARY 1962.

Richardson numbers calculated from Equations (1), (9), and (13) were comparable; significantly different values of Ri were obtained using Eq. (12). The Bernstein and Young (1962) gradient corrections were applied to calculations for Eq. (1) only. The stability ratio calculations showed no significant differences between Eq. (14) and Eq. (15). The departures observed using Eq. (12) are attributed to certain inaccuracies in obtaining log θ' owing to values of log $\theta_2-\theta_1/\log (z_1 - z_2)$ being very close to zero. All results are considered to be within the limits of accuracy of the instruments used to obtain the data.

Absolute values of Richardson numbers with respect to height above the surface are presented to Table V. One hundred forty-seven wind and temperature profiles of data obtained from the research tower were used in this phase of

Height (Meters)	Stable	Unstable	Extremely Unstable
	Ri	Ri	Ri
2.95	0.113	0.026	0.360
5.33	0.309	0.056	0.501
9.65	0.610	0.063	0.326
15.92	0.812	0.193	1.290
22.10	1.610	0.276	1.836
28.22	0.535	0.356	1.903
34.36	0.882	0.484	2.930
43.21	1.287	0.687	3.199
55.50	1.176	1.098	7.162

Table V. Absolute values of the gradient Richardson number for the first 62 meters of the boundary layer.

the study. The profiles were classified as stable, unstable, and extremely unstable using the Richardson number at 2.95 meters as the classifier. The stable cases included all values of $Ri > 0$, while the unstable cases were in the range $0 > Ri > -0.05$, and the extremely unstable regime was $Ri < -0.05$. From Table V it will be seen that only the stable case provides values of Ri that are comparable to those listed in Table I, indicating that nocturnal stability is similar in two distinctly separate locales. On the other hand, the daytime cases represented by the unstable and extremely unstable data show that stability is a function of insolation, ground cover, fetch, and roughness discontinuities. The data in Table I were obtained during the Great Plains Turbulence Program at O'Neill, Nebraska, over reasonably homogeneous terrain, while the data from WSMR were obtained over heterogeneous terrain in a semi-arid region. Apparently, a combination of transparency of the atmosphere, surface roughness, nonuniform terrain and nonstationary flow conditions lead to atmospheric processes that are basically more unstable than observed over the Great Plains. Local influences at the WSMR observational site include the formation of internal momentum and thermal boundaries, local advection of heat and momentum, and conditions of windless convection as defined by Lumley and Panofsky (1964).

The Richardson numbers determined for the tower locale generally increase in absolute magnitude with height, with some discontinuities in the general trend. For the stable case, the abrupt increase in Ri between 15.92 and 22.10 meters is indicative of advection of heat and adiabatic warming of the atmosphere under nocturnal drainage wind conditions. The unstable and extremely unstable cases also reflect advection of heat and particularly the internal thermal boundary formation as indicated by the values of Ri at 9.65 and 15.92 meters. Windless convection conditions are apparent in the extremely unstable cases, from the large values of Ri observed above 15.92 meters.

CONCLUSIONS

The accurate determination of the Richardson number for micrometeorological purposes is highly dependent upon proper evaluation of the vertical gradients of wind and potential temperature in the first few meters of the atmosphere. The presence of heterogeneous processes in the planetary boundary layer leads to improper evaluation of the vertical gradients if these phenomena are not recognized and compensated for in the analysis of the data.

The existence of a gap in the wind speed spectrum with a period of approximately one hour separating the micro- from the mesoscale processes in the boundary layer indicates that commensurate averaging times are needed to provide adequate information on the stability of the lowest few meters of the atmosphere.

Values of Ri for the first 62 meters above the surface as computed for the White Sands Research Tower are considered to be representative for the locale and the flow conditions prevalent when the experimental data were obtained. Failure to take into account the terrain over which the flow occurs or the turbulent processes in operation could lead to erroneous evaluation of the stability regime or rejection of the data as unrealistic.

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13 ABSTRACT

Determination of the stability regime is a basic approach in any investigation of atmospheric turbulence. The establishment of stability criteria in the boundary layer is usually accomplished by use of the nondimensional Richardson number. The computation of accurate Richardson numbers is shown to be adversely affected by a number of factors including the choice of vertical gradients, the terrain, spacing of instruments, and heterogeneous profiles of wind and temperature.

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